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Estimated Hydraulic Properties for the Surficial- and Bedrock-Aquifer System, Meddybemps, Maine

Open-File Report 99-199

Prepared in cooperation with the
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1. REPORT DATE	2. REPORT TYPE	3. DATES COVERED
1999	**N/A**	**-**

4. TITLE AND SUBTITLE **Estimated Hydraulic Properties for the Surficial - and Bedrock-Aquifer System, Meddybemps, Maine**	5a. CONTRACT NUMBER 5b. GRANT NUMBER 5c. PROGRAM ELEMENT NUMBER
6. AUTHOR(S)	5d. PROJECT NUMBER 5e. TASK NUMBER 5f. WORK UNIT NUMBER
7. PERFORMING ORGANIZATION NAME(S) AND ADDRESS(ES) **U.S. Department of the Interior 1849 C Street, NW Washington, DC 20240**	8. PERFORMING ORGANIZATION REPORT NUMBER
9. SPONSORING/MONITORING AGENCY NAME(S) AND ADDRESS(ES)	10. SPONSOR/MONITOR'S ACRONYM(S) 11. SPONSOR/MONITOR'S REPORT NUMBER(S)

12. DISTRIBUTION/AVAILABILITY STATEMENT
Approved for public release, distribution unlimited

13. SUPPLEMENTARY NOTES
The original document contains color images.

14. ABSTRACT

15. SUBJECT TERMS

16. SECURITY CLASSIFICATION OF:			17. LIMITATION OF ABSTRACT	18. NUMBER OF PAGES	19a. NAME OF RESPONSIBLE PERSON
a. REPORT	b. ABSTRACT	c. THIS PAGE			
unclassified	**unclassified**	**unclassified**	**UU**	**32**	

Standard Form 298 (Rev. 8-98)
Prescribed by ANSI Std Z39-18

U.S. Department of the Interior
U.S. Geological Survey

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By FOREST P. LYFORD, STEPHEN P. GARABEDIAN, and BRUCE P. HANSEN

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Northborough, Massachusetts
1999

U.S. DEPARTMENT OF THE INTERIOR
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CONVERSION FACTORS

Multiply	By	To obtain
acre	0.4047	hectare
cubic foot per day (ft^3/d)	28.317	liters per day
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot per minute (ft/min)	0.3048	meter per minute
foot squared per day (ft^2/d)	0.09290	meter squared per day
gallons per day (gal/d)	0.003785	cubic meter per day
gallon per minute (gal/min)	0.06309	liter per second
gallon per minute per foot [(gal/min)/ft)]	0.2070	liter per second per meter
inch (in.)	25.4	millimeter
inch per year (in/yr)	25.4	millimeter per year
mile (mi)	1.609	kilometer

VERTICAL DATUM

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)--a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

Estimated Hydraulic Properties for the Surficial- and Bedrock-Aquifer System, Meddybemps, Maine

By Forest P. Lyford, Stephen P. Garabedian, *and* Bruce P. Hansen

Abstract

Analytical and numerical-modeling methods were used to estimate hydraulic properties of the aquifer system underlying the Eastern Surplus Company Superfund Site in Meddybemps, Maine. Estimates of hydraulic properties are needed to evaluate pathways for contaminants in ground water and to support evaluation and selection of remediation measures for contaminated ground water at this site.

The hydraulic conductivity of surficial materials, determined from specific-capacity tests, ranges from 17 to 78 feet per day for wells completed in coarse-grained glaciomarine sediments, and from about 0.1 to 1.0 foot per day for wells completed in till. The transmissivity of fractured bedrock determined from specific-capacity tests and aquifer tests in wells completed in less than 200 feet of bedrock ranges from about 0.09 to 130 feet squared per day. Relatively high values of transmissivity at the south end of the study area appear to be associated with a high-angle fracture or fracture zone that hydraulically connects two wells completed in bedrock. Transmissivities at six low-yielding (less than 0.5 gallon per minute) wells, which appear to lie within a poorly transmissive block of the bedrock, are consistently in a range of about 0.09 to 0.5 foot squared per day.

The estimates of hydraulic conductivity and transmissivity in the southern half of the study area are supported by results of steady-state calibration of a numerical model and simulation of a 24-hour pumping test at a well completed in bedrock. Hydraulic conductivity values for the surficial aquifer used in the model were 30 feet per day for coarse-grained glaciomarine sediments, 0.001 to 0.01 foot per day for fine-grained glaciomarine sediments, and 0.1 to 0.5 foot per day for till. As part of model calibration, a relatively transmissive zone in the surficial aquifer was extended beyond the hypothesized extent of coarse-grained sediments eastward to the Dennys River.

Hydraulic conductivity values used for bedrock in the model ranged from 3×10^{-4} to 1.5 feet per day. The highest values were in the fracture zone that hydraulically connects two wells and apparently extends to the Dennys River. The transmissivity of bedrock used in the model ranged from 0.03 to 150 feet squared per day, with the majority of the bedrock transmissivities set at 0.3 foot squared per day. Numerical modeling results indicated that a very low ratio of vertical hydraulic conductivity to thickness (1×10^{-9} days^{-1}) was required to simulate a persistent cone of depression near a residential well that lies in the previously identified poorly transmissive block of bedrock.

INTRODUCTION

Volatile organic compounds (VOCs) have been detected in ground water in surficial materials and bedrock in two areas near the Eastern Surplus Superfund Site in Meddybemps, Maine (Lyford and others, 1998). The U.S. Environmental Protection Agency (USEPA) and local residents are concerned that contaminants, principally tetrachloroethylene (PCE), can move to existing residential wells and limit future development of ground-water resources in the area. Information on the hydraulic properties of the aquifer system is needed to assess the potential for contaminants in ground water to affect residential wells

and to help evaluate remediation approaches. During 1997–98, the U.S. Geological Survey (USGS), in cooperation with USEPA, studied characteristics of the fractured crystalline bedrock and the hydraulic properties of the aquifer system near the Eastern Surplus Superfund Site.

The purpose of this report is to provide estimates of hydraulic properties for the surficial- and bedrock-aquifer system near the Eastern Surplus Superfund Site. Estimates of transmissivity and hydraulic conductivity determined from specific-capacity data and aquifer tests were refined by calibration of a numerical model for steady-state and transient conditions. Calibration of the numerical model also provided estimates of vertical hydraulic conductivity and reinforced the conceptual ground-water-flow model for the study area. The study focused on the area near the southernmost of two contaminant plumes that is closest to existing residential wells. The characteristics of fractures near the site are described in a companion report (Hansen and others, 1999).

Thanks are extended to Edward Hathaway, USEPA Project Manager, for logistical support during the study and to Mona Van Wart and Charlotte Smith for access to their wells during aquifer testing. Also, thanks are extended to Madge Orchard, Terry Lord, Greg Smith, and Harry Smith for access to their property.

DESCRIPTION OF STUDY AREA

The Eastern Surplus Superfund Site is on the western bank of the Dennys River at the outlet from Meddybemps Lake (figs. 1 and 2). A study area of approximately 30 acres encompasses the 4-acre Superfund site. The primary focus of this investigation was on the southern half of the study area. The following description of the hydrogeology of the region and study area is summarized from Lyford and others (1998).

The region that encompasses the study area is underlain mainly by the Meddybemps Granite. A small area centered on the study area is underlain by a gabbro-diorite, which is most likely a detached body of mafic rock within the Meddybemps Granite. Surficial materials include till, generally less than 10 to 20 ft thick, and extensive glaciomarine deposits, including coarse-grained and fine-grained (Presumpscot Formation) sediments deposited during deglaciation of the region. The coarse gravel and sand was deposited in an ancestral sea, probably as a subaqueous fan at the ice margin during retreat of the glacier. Glaciomarine silty clay of the Presumpscot Formation underlies much of the lowland area in the region. A sandy facies in the upper section of the Presumpscot Formation was deposited as the land rose relative to sea level and the shoreline regressed southeastward through the area.

Hydrogeologic units in the study area include till, coarse-grained glaciomarine deposits, fine-grained glaciomarine deposits, and bedrock. The vertical and lateral distribution of hydrogeologic units is shown in sections on figure 3. Till thickness ranges from less than 5 ft on the western side of the Dennys River to about 40 ft on the eastern side. The coarse-grained glaciomarine deposits are present at or near the surface in the western part of the study area; thickness ranges from 0 to more than 30 ft. The thick (more than 10 ft) coarse-grained sections are largely above the water table. Fine-grained glaciomarine deposits (Presumpscot Formation) are present in the central and southern parts of the study area where thickness ranges from 0 to about 20 ft. The silt-clay facies of this unit is poorly permeable and serves as a confining layer for ground water in underlying till and coarse-grained glaciomarine deposits. Ground water in bedrock occurs principally in fractures. The occurrence of water-yielding fractures ranges widely; in some wells only one or two fractures supply measurable quantities of water (more than 0.02 gal/min).

Ground-water levels in bedrock wells on the north side of the study area respond rapidly to rainfall. Responses to precipitation in surficial materials and bedrock are subdued or are not apparent where silts and clays of the Presumpscot Formation are present. The annual recharge may approach a potential rate of 24 to 26 in. where coarse-grained materials are present at the surface, but is probably less where till, silts, and clays are at the surface.

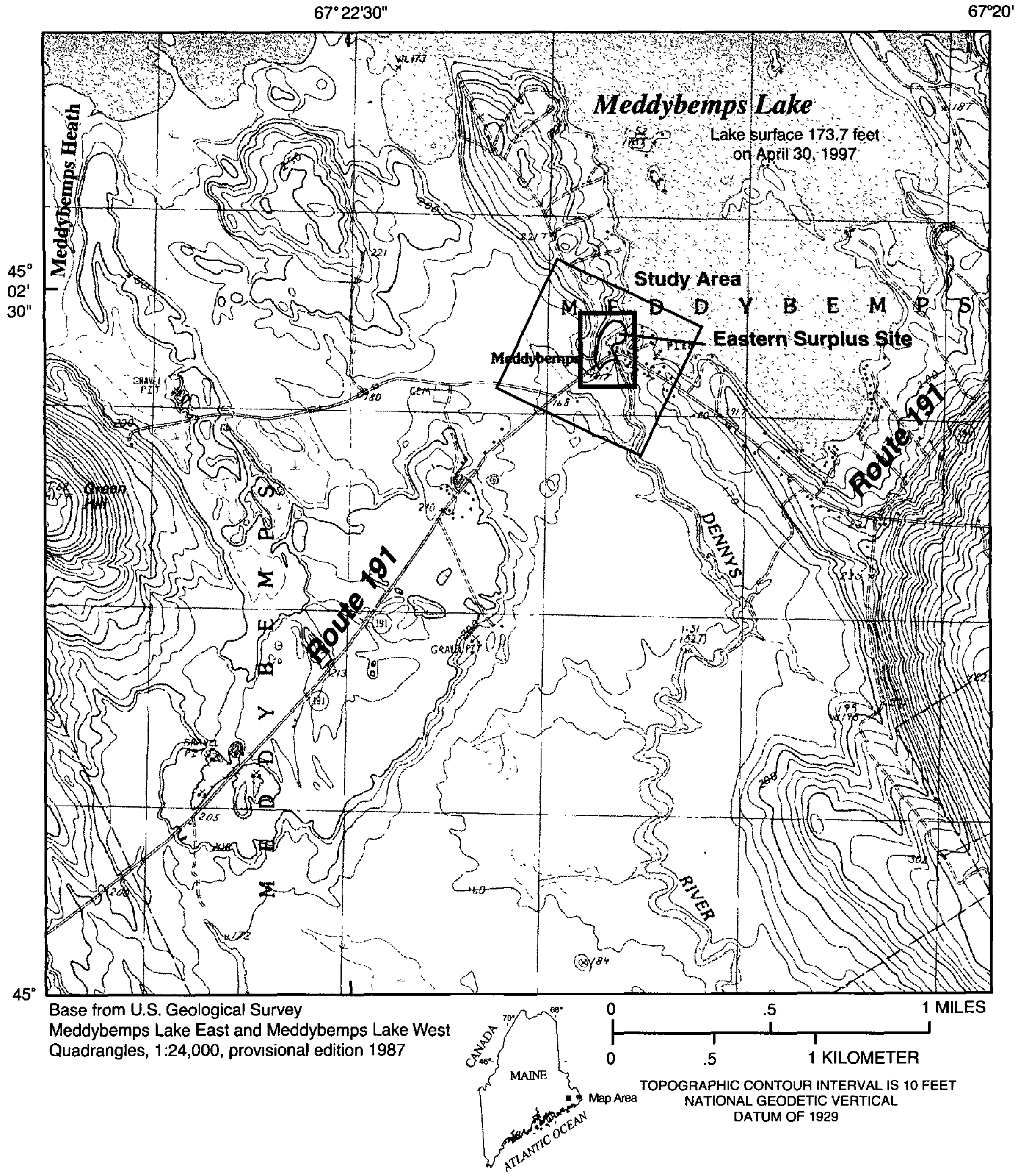


Figure 1. Location of the Eastern Surplus Superfund Site, study area, and numerical model area, Meddybemps, Maine. (Modified from Lyford and others, 1998, fig. 1.)

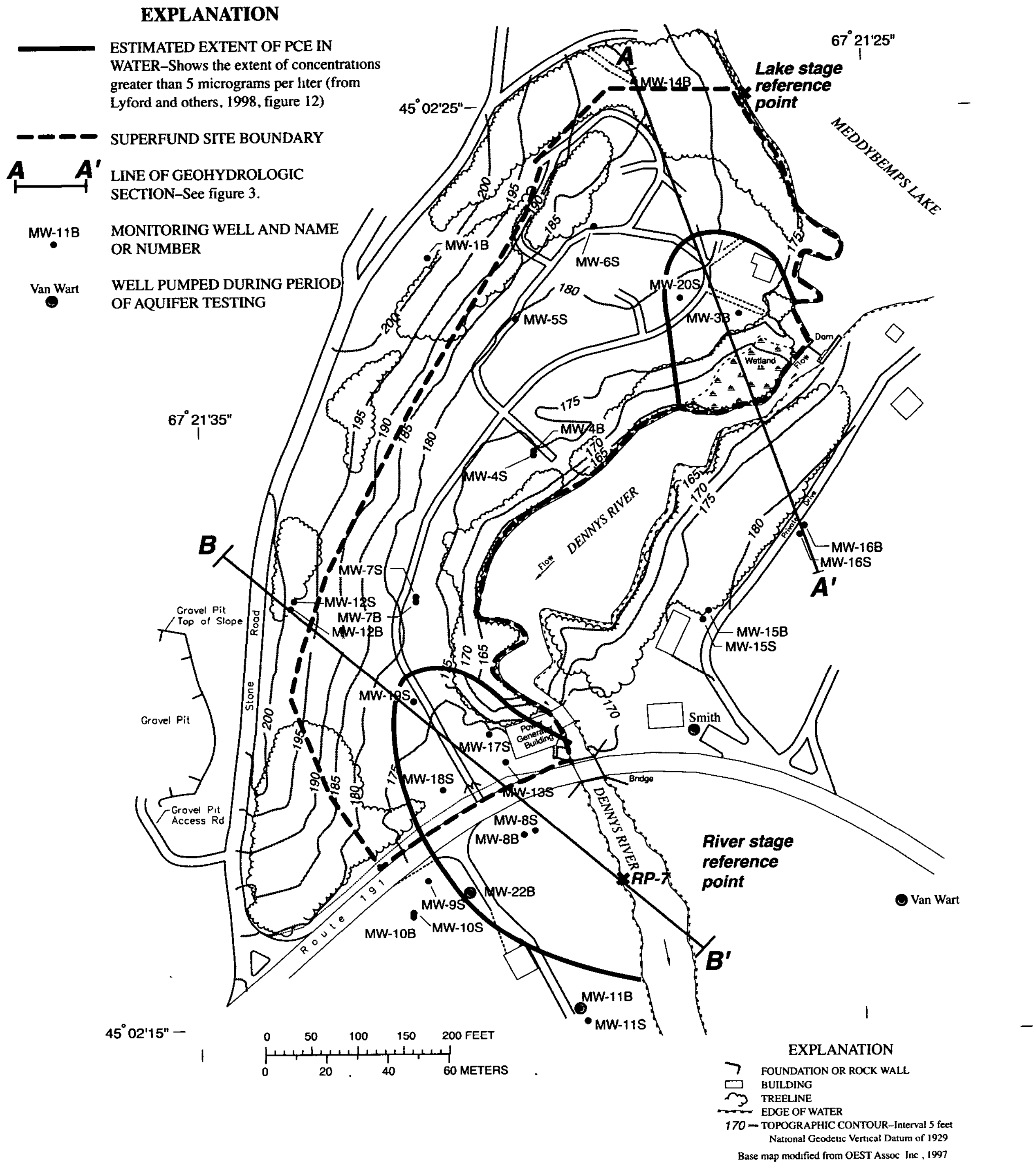


Figure 2. Location of study area, extent of tetrachloroethylene (PCE) in ground water and locations of wells, Meddybemps, Maine.

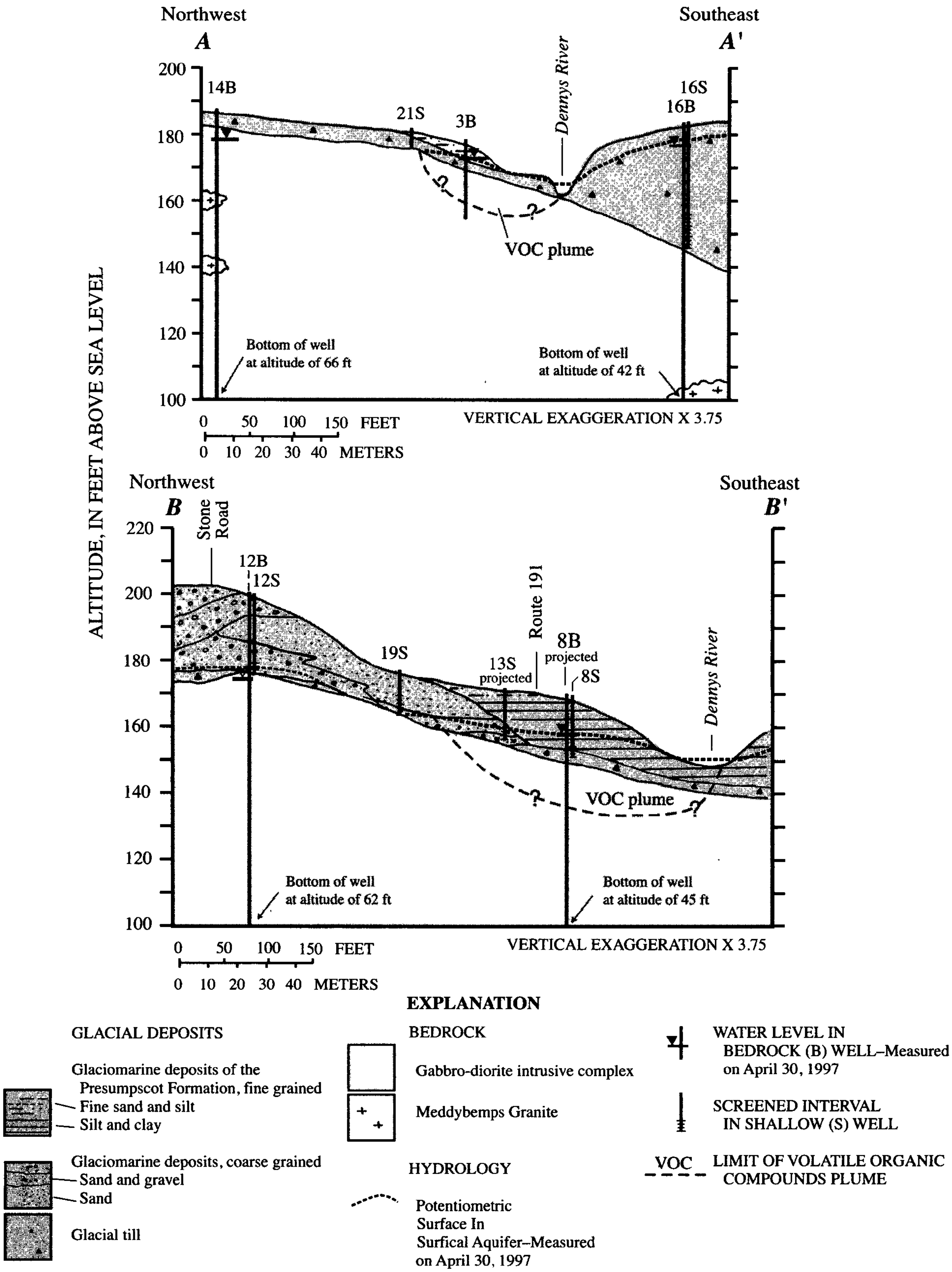


Figure 3. Geohydrologic sections A-A′ and B-B′, Meddybemps, Maine.

Ground water in surficial materials generally flows towards the Dennys River. The saturated thickness of surficial materials under the study area is generally less than 10 ft, and several monitoring wells screened in surficial materials "go dry" during extended periods of little or no recharge. South of Highway 191, ground water in coarse-grained sediments is confined below silts and clays of the Presumpscot Formation, but flow also is toward the Dennys River.

Ground water in bedrock flows towards the Dennys River from the eastern and western sides of the study area. In addition, hydraulic gradients are generally downward from the surficial aquifer to bedrock. Water-level data also indicate a potential for flow under the river from the western side to a cone of depression near a residential well on the eastern side. The cone of depression may extend laterally several hundred feet and affect water levels within a block of the aquifer characterized by low-yielding wells (less than 0.5 gal/min). In contrast, it is likely that the higher-yielding bedrock wells outside the low-yielding zone intersect high-angle fractures that extend to the overlying surficial aquifer.

Plumes of VOCs, including PCE and trichloroethylene (TCE), have been detected in ground water in two areas. VOCs in both plumes move through surficial materials and shallow bedrock towards the Dennys River. Contaminants in the southern plume could potentially move through fractures in bedrock to the local cone of depression east of the Dennys River.

ESTIMATION OF HYDRAULIC PROPERTIES USING ANALYTICAL METHODS

Specific-capacity data and drawdown data from aquifer tests conducted in wells in the study area were used to estimate aquifer transmissivity and hydraulic conductivity by applying conventional analytical methods. For tests in bedrock, an equivalent porous medium and radial flow are assumed. This section describes results of these analyses.

Specific-Capacity Measurements

The specific capacity of a well is the ratio of pumping rate of the well to drawdown at the well. The following formula presented by Cooper and Jacob (1946) is used here to estimate aquifer transmissivity from specific capacity data. Application of the method has been described by Fetter (1994).

$$T = \frac{2.30Q}{4\pi s}\log\frac{2.25Tt}{r^2S}, \tag{1}$$

where:

T = transmissivity,
s = drawdown,
Q = pumping rate,
S = storage coefficient,
t = time since pumping began, and
r = well radius.

Application of equation 1 requires an estimate of the storage coefficient and an initial estimate of transmissivity. The final transmissivity is then determined through a series of iterations that uses the previously estimated or calculated value of transmissivity. An initial estimate of transmissivity, in feet squared per day, can be derived by multiplying the specific capacity of the well, in gallons per minute per foot of drawdown, times 200. The transmissivity determined by the specific-capacity method is, however, subject to considerable uncertainty. In addition to using an estimate of storage coefficient, which could be in error by a factor of 10 or more, the method requires the following assumptions: (1) well-entry losses are negligible, (2) the pumping period is sufficiently long to satisfy the requirements for applying the Cooper-Jacob formula ($r^2S/4Tt < 0.01$), and (3) wellbore storage effects are negligible. At a minimum, transmissivity values determined from specific-capacity tests provide relative values that can be used as indices for some types of hydrologic analyses. Specific-capacity data were collected during water-quality sampling (Roy F. Weston, Inc., 1997; 1998a), during borehole flowmeter tests, during recovery measurements in bedrock wells after drilling, and during aquifer tests. Well-construction data for the wells tested are summarized in table 1.

Table 1. Well construction data, water levels and stage on September 9–10, 1997, and wells equipped with water-level recorders during aquifer testing, Meddybemps, Maine

[All depths in feet below land surface except water level and stage, which are in feet below the measuring point. ft, feet; No., number; --, no data; na, not applicable]

Well or reference point No. or name	Date drilled	Altitude of land surface (ft)	Altitude of measuring point (ft)	Total depth of well (ft)	Depth to bedrock or refusal (r) (ft)	Screened (s) or open-hole (o) interval (ft)	Water level or stage below measuring point (ft)	Water-level altitude (ft)	Equipped with recorder during aquifer testing
MW-1B	4/17/88	201.60	204.18	57.8	34.6	s 38–53	36.62	167.56	no
MW-3B	4/17/88	177.37	179.89	23.3	9	s 13.3–23.3	8.96	170.93	no
MW-4S	4/15/88	174.84	177.60	18	19.5	s 13.0–18.0	13.82	163.78	no
MW-4B	4/14/88	174.75	177.43	39.7	19.5	s 24.7–39.7	15.48	161.95	no
MW-5S	10/23/96	179.86	182.06	13	r 13	s 10–13	13.78	168.28	no
MW-6S	10/23/96	182.34	184.71	7.0	r 7.0	o 4.5–7.0	dry	--	no
MW-7S	10/28/96	177.79	180.09	17.2	17	s 12–17	18.22	161.57	no
MW-7B	10/28/96	177.81	178.75	117.8	18	o 21–117.8	22.12	156.63	yes
MW-8S	10/25/96	167.30	169.14	16.5	r 16.5	s 14–16.5	12.66	156.48	yes
MW-8B	11/04/96	169.04	169.35	124	20.5	o 25.7–124	13.45	155.90	yes
MW-9S	10/25/96	174.03	175.52	16.5	r 16.5	s 14–16.5	dry	--	no
MW-10S	11/06/96	174.42	176.13	23	22	s 18–23	18.58	157.55	yes
MW-10B	11/4/96	174.24	175.64	120	20	o 26.4–120	18.97	156.67	yes
MW-11S	10/26/96	169.34	170.70	26	r 26	s 21–26	16.55	154.15	yes
MW-11B	11/04/96	169.69	170.63	129	29	o 35.1–129	15.83	154.80	yes
MW-12S	10/26/96	199.11	200.21	22	r 22	s 19–21.5	dry	--	no
MW-12B	11/4/96	200.13	201.34	138	22.5	o 27.7–138	26.72	174.62	yes
MW-13S	10/29/96	171.36	174.14	14	r 14	s 11–13.5	dry	--	no
MW-14B	11/05/96	185.70	187.33	120	3.5	o 9.4–120	14.17	173.16	no
MW-15S	11/06/96	178.46	179.32	38	36	s 26–36	17.08	162.24	no
MW-15B	11/05/96	178.97	180.11	240	39	o 46.9–240	25.76	154.35	yes
MW-16S	11/06/96	182.88	183.48	38	36	s 28–38	12.60	170.88	yes
MW-16B	11/05/96	182.18	183.91	138	38	o 42.3–140	13.39	170.52	yes
MW-17S	4/22/97	172.42	174.34	23	18.0	s 15–17.5	15.98	158.47	yes
MW-18S	4/23/97	172.90	174.82	19.5	18.0	s 16–18.5	17.10	157.85	yes
MW-19S	4/23/97	177.08	178.60	13.5	11.8	s 9.3–11.8	dry	--	no
MW-20S	4/24/97	178.57	180.33	8.0	6.0	s 3.5–6.0	dry	--	no
MW-22B	1950s	172.35	174.28	49	18	o 25.5–49	17.47	156.81	yes
VanWart	--	171.78	174.13	142	29	o 39–142	9.15	164.98	yes
Smith	--	173.35	174.55	420	--	--	110.10	[1]64.45	yes
Meddybemps Lake	na	na	174.09	na	na	na	2.33	171.76	no
Dennys River at RP-7	na	na	156.41	na	na	na	3.85	152.56	no

[1]Water level in the Smith well was recovering slowly at the time of measurement.

During sampling for water quality, wells were pumped at nearly constant rates that resulted in minimal and nearly constant drawdown. Pumping continued until water-quality parameters measured in the field had stabilized. Data are available for sampling periods in December 1996, June 1997, and October 1997. The pumping rate differed with well yield and was typically less than 0.4 gal/min. Yields for wells MW-11S, MW-4B, MW-8B, MW-12B, MW-15B, and MW-16B (fig. 1) were too low to sustain a constant pumping rate. These wells were sampled after purging and allowing the water level to recover (Roy F. Weston, Inc., 1998a). Several wells were dry or nearly dry in October 1997 after an extended period of low precipitation and could not be sampled. Specific-capacity data also were collected during borehole-flowmeter logging at the Van Wart well, during a brief (30 min) aquifer test at well MW-3B, and during aquifer tests at wells MW-11B and MW-22B.

A storage coefficient of 0.1 was assumed for most wells completed in the surficial aquifer. This value is within the range commonly assumed for unconfined aquifers (Lohman, 1972, p. 8) and is considered to be a reasonable estimate for the short-term tests that were typically 45 to 90 min in duration. Exceptions were wells MW-8S and MW-10S, where ground water is confined by clays of the Presumpscot Formation, and wells MW-15S and MW-16S, where the well screen is considerably below the water table. A storage coefficient of 1×10^{-4}, which is within the range commonly assumed for confined aquifers (Lohman, 1972, p. 8), was assumed for these wells. A storage coefficient of 1×10^{-4} was also assumed for all wells completed in bedrock. This value is at the high end of a range from 5×10^{-7} to 1×10^{-4} reported for fractured-rock aquifers (Earl Greene, U.S. Geological Survey, written commun., 1997). The importance of the storage-coefficient estimates will be discussed later in this section of the report.

An alternative approach to pumping was used to determine the specific capacity of bedrock wells MW-8B, MW-12B, MW-15B, and MW-16B. Water levels in these wells recovered slowly and at a nearly constant rate, indicating a constant inflow rate, for several hours to several days after they were drilled (fig. 4). Conceptually, when the water level in a well declines below the level of a bedrock fracture, drawdown at the well bore is effectively constant and flow to the well would be expected to gradually decline (Jacob and Lohman, 1952). A steady inflow rate while water-bearing fractures are above the water level in the well indicates either that a constant head source, such as leakage from a surficial aquifer, was controlling the inflow rate, or that sufficient time had elapsed so changes in discharge were very slow and not discernible in the recovery record. Computation of changes in flow with time using an equation presented by Jacob and Lohman (1952), and estimates of hydraulic properties at these low-yield wells, indicated that changes in flow should have been discernible during the recovery period. For this reason, leakage from a nearby source, causing steady flow to a well after a relatively short time, seems to be the more likely cause of the steady inflow rate. For this analysis, a time of 100 min for flow to stabilize was assumed.

A transmissivity value was computed for major fractures or fracture zones in these four bedrock wells using equation 1. Equation 1 was derived for constant-flow conditions, but it also applies to constant-drawdown conditions after very small values of time (Lohman, 1972, p. 23). The well yield, in cubic feet per day (ft^3/d), was computed by multiplying the rate of water-level rise by the cross-sectional area of the well. The yield for each fracture was assumed to be the percentage of total yield determined from borehole-flowmeter tests (Hansen and others, 1999). The drawdown for each fracture or fracture zone was the depth to the fracture below a static water level that was measured about 2 weeks after the well was drilled. The depths to water-bearing fractures and percentage of flow reported by Hansen and others (1999) are shown in figure 4. The transmissivity values reported in table 2 are the sum of transmissivity values computed for all water-bearing fractures or fracture zones in a well.

The estimates of transmissivity for all the bedrock wells that were made on the basis of specific-capacity data range widely—from 0.09 ft^2/d in well MW-15B to 130 ft^2/d in well MW-3B. The transmissivity range of 280 to 550 ft^2/d for well MW-22B shown in table 2 probably reflects the transmissivity of the bedrock and surficial aquifers combined, as discussed later in this report. For this reason, the estimates of transmissivity for well MW-22B are not included in this range. The transmissivity values for the low water-yielding wells MW-7B, MW-8B, MW-12B, MW-15B, MW-16B, and possibly MW-4B, are 0.5 ft^2/d or less. For comparison, Lyford and others (1998) report a transmissivity of 0.6 ft^2/d for the Smith Well, another low-yielding well.

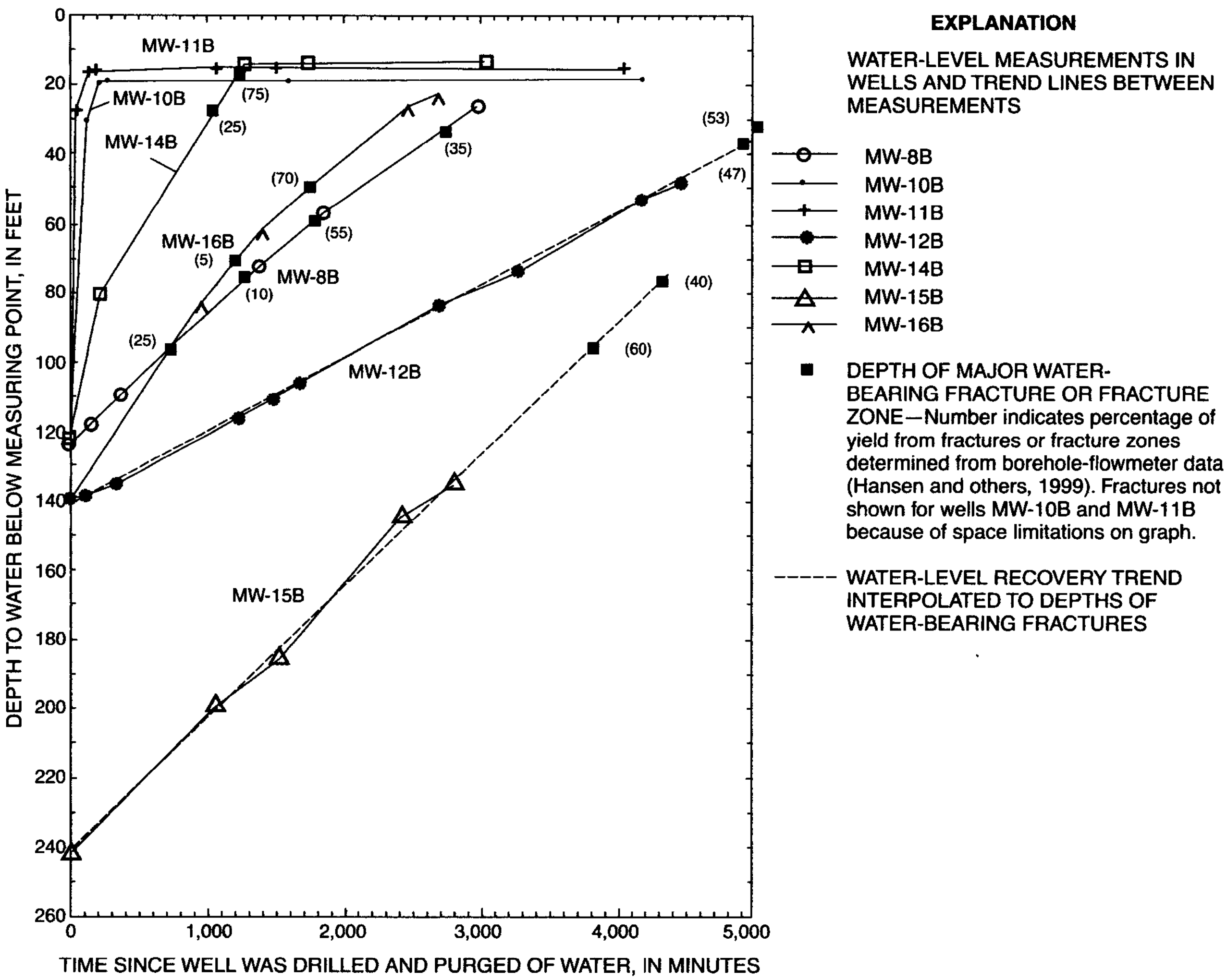


Figure 4. Depth to water during recovery of water levels in selected bedrock wells after drilling and depths of water-bearing fractures or fracture zones.

The relatively high transmissivity values determined for wells MW-10B, MW-11B, MW-14B, and the Van Wart well, can be attributed to the presence of one or more water-yielding fractures or fracture zones (Hansen and others, 1999). The high transmissivity measured in well MW-3B may be attributable to fractures in shallow bedrock that provide a hydraulic connection to the surficial aquifer and to Meddybemps Lake, but fracture data were not available for this well.

Estimates of transmissivity and hydraulic conductivity for surficial materials that were made on the basis of specific-capacity data are summarized in table 3. The hydraulic conductivity values reported in table 3 were determined by dividing the transmissivity of each well by the saturated thickness of the aquifer at the time of the tests. This approach for estimating hydraulic conductivity requires the assumption that the full saturated thickness of the aquifer contributed water to the well during pumping. This assumption is reasonable if the saturated thickness is approximately the same as the length of the well screen, and for wells completed in coarse-grained materials. The assumption may yield values of hydraulic conductivity that are somewhat lower than actual values in fine-grained materials if vertical hydraulic conductivities are relatively low, and if the screen length is considerably less than the saturated thickness. The hydraulic conductivity estimates in table 3 for wells MW-8S, MW-15S, and MW-16S may be lower than actual values for this reason.

Table 2. Estimates of transmissivity for bedrock using specific-capacity data for wells, Meddybemps, Maine

[o, length of open borehole; s, length of screen; ft^2/d, feet squared per day; ft, feet]

Well name	Transmissivity range (ft^2/d)	Number of tests	Length of screen or open borehole (ft)
MW-1B	1.3–2.3	2	s 15
MW-3B	60–130	4	s 10
MW-4B	<1.4	1	o 15
MW-7B	0.1–0.5	2	o 97
MW-8B	0.2	1	o 98
MW-10B	8.0–15	3	o 94
MW-11B	70–110	4	o 94
MW-12B	0.3	1	o 110
MW-14B	20–32	3	o 111
MW-15B	0.09	1	o 193
MW-16B	0.2	1	o 98
MW-22B	280–550	2	o 23
Van Wart	12	1	o 103

Table 3. Estimates of transmissivity and hydraulic conductivity for surficial materials using specific-capacity data for wells, Meddybemps, Maine

[ft, foot; ft/d, feet per day; ft^2/d, feet squared per day]

Well name	Transmissivity range (ft^2/d)	Number of tests	Screen length (ft)	Saturated thickness range (ft)	Hydraulic conductivity range (ft/d)
MW-4S	1.5–5.2	3	5.0	5.5–9.2	0.3–0.6
MW-5S	150–200	2	3.0	6.2–8.6	17–32
MW-7S	230–290	2	5.0	3.7–3.7	63–78
MW-8S	3.9	1	2.5	7.4	0.5
MW-10S	210–220	2	5	5.0–7.2	29–45
MW-13S	50	1	2.5	2.7	19
MW-15S	7.6–19	3	10	18.4–23.7	0.3–1.0
MW-16S	2.9–8.0	3	10	23.1–30.6	0.1–0.3
MW-17S	11–18	2	2.5	3.6–6.7	2.7–3.2
MW-18S	120	1	2.5	5.7	20

Relatively high values of transmissivity and hydraulic conductivity for surficial materials were measured in wells MW-5S, MW-7S, MW-10S, and MW-18S. Estimates of hydraulic conductivity that range from 17 to 78 ft/d in these wells reflect the likely range of hydraulic properties for coarse-grained glaciomarine sediments and is characteristic of silty sand and clean sand (Freeze and Cherry, 1979, table 2.2). Hydraulic conductivity values of 0.1 to 1 ft/d in wells MW-4S, MW-8S, MW-15S, and MW-16S reflect the likely range of hydraulic conductivity for till. This range is within the range of values for tills derived from crystalline rocks in southern New England and northern New Hampshire (Torak, 1979: Pietras, 1981; Melvin and others, 1992, table 3; Tiedeman and others, 1997, p.8).

Specific-capacity data are not available for wells MW-9S, MW-12S, MW-19S, and MW-20S because they were dry or nearly dry and were not sampled. Specific-capacity data are not available for well MW-11S because the yield was too low to sustain pumping during sampling.

The major uncertainty associated with the specific-capacity approach for estimating aquifer transmissivity is the storage coefficient. The computed values of transmissivity, however, are relatively insensitive to the storage coefficient because the coefficient appears in the log term of equation 1. For example, for well MW-12B completed in bedrock, a reduction of the storage coefficient from 0.0001 to 0.00001 increases the transmissivity from 0.16 to 0.21 ft^2/d. For well MW-7S in the surficial aquifer, increasing the storage coefficient from 0.1 to 0.2 decreases the transmissivity from 230 ft^2/d to 207 ft^2/d, and decreasing the storage coefficient from 0.1 to 0.05 increases the transmissivity to 250 ft^2/d. For the low-yielding wells, the time required for the inflow rate to stabilize also is uncertain, but, as with the storage coefficient, the transmissivity estimates are not particularly sensitive to time because time also appears in the log term of equation 1. For example, for well MW-12B, a reduction of the time from 100 min to 10 min in equation 1 reduces the computed transmissivity of the highest-yielding fracture from 0.16 ft^2/d to 0.11 ft^2/d.

Aquifer Tests

Aquifer tests were conducted at wells MW-22B and MW-11B during September 10–14, 1997, to refine estimates of hydraulic properties for the ground-water system. Water-level responses to pumping were observed in these two wells and the Smith and Van Wart wells, which were pumped intermittently for domestic purposes during the aquifer-test period (fig. 5). Water levels in wells that responded to pumping are shown in figure 6. Water levels in the Dennys River, Meddybemps Lake, and wells prior to aquifer testing are summarized in table 1.

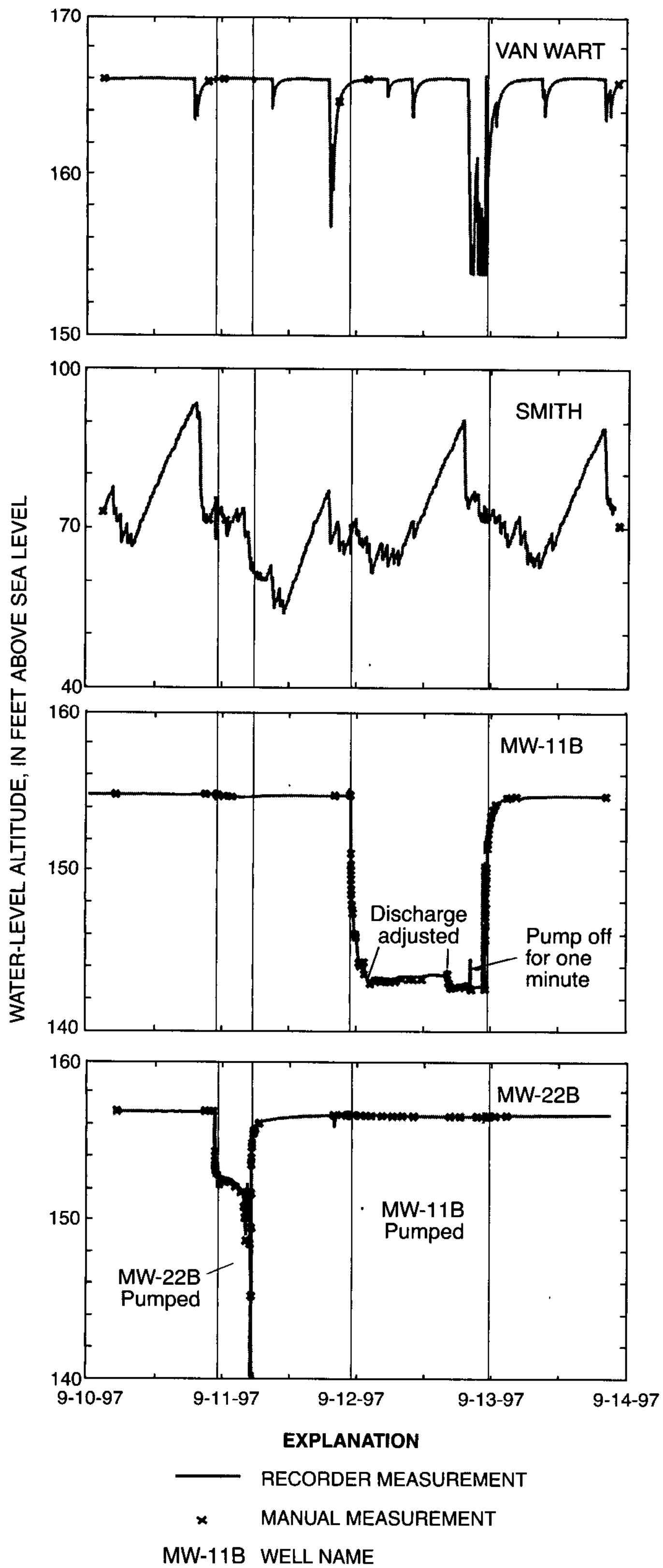


Figure 5. Water-level hydrographs for wells pumped during aquifer testing, Meddybemps, Maine.

The original plan for aquifer testing was to pump well MW-22B at a constant rate for at least 24 hours. After about 6 hours of pumping at a rate of 9 gal/min, however, drawdown suddenly increased and the water level quickly fell to the pump intake. During part of the pumping period, turbidity in the water caused by fine sand and silt indicated that the well may not be fully cased through surficial materials or that fractures in bedrock provided a short connection to the surficial aquifer. The test was terminated after 325 minutes, and water levels were allowed to recover for about 18 hours before a second test was started in well MW-11B.

Water-level data for the first 150 minutes of the test in well MW-22B were used to estimate aquifer transmissivity by applying the straight-line method of Cooper and Jacob (1946). Drawdown during this part of the test followed an approximate straight-line trend on a semi-logarithmic plot, as shown in figure 7, before the well started producing sand and silt and the water level started declining at a greater rate. A transmissivity value of 450 ft^2/d estimated from the time and drawdown data is consistent with estimates using specific-capacity data (table 2). This estimate of transmissivity may be high relative to other wells because it may represent a combination of the surficial and bedrock aquifers.

For the second test, well MW-11B was pumped at a rate of 4.5 gal/min for 24 hours. A uniform pumping rate was difficult to maintain, and the small variations in pumping rate were apparent in the water-level record. Water levels in all wells monitored south of Route 191, except well MW-8B, responded to pumping from well MW-11B (fig. 5). A rise of water level in well MW-8B shortly after pumping started resulted from an estimated 0.1 to 0.2 in. of precipitation that entered the uncovered well during a rain shower.

Water-level data for the first 200 min of the test in well MW-11B (fig. 7) were used to estimate transmissivity by applying the straight-line method of Cooper and Jacob (1946). During this period, the effects of leakage and well-bore storage were assumed to be negligible; however, both factors could have affected the analysis. A transmissivity of 38 ft^2/d calculated by this method is lower than estimates using specific-capacity data (70–110 ft^2/d in table 2), possibly because well-bore storage affected the rate of drawdown during the first hour of pumping. The estimates of transmissivity for this well were refined using numerical methods discussed in the next section.

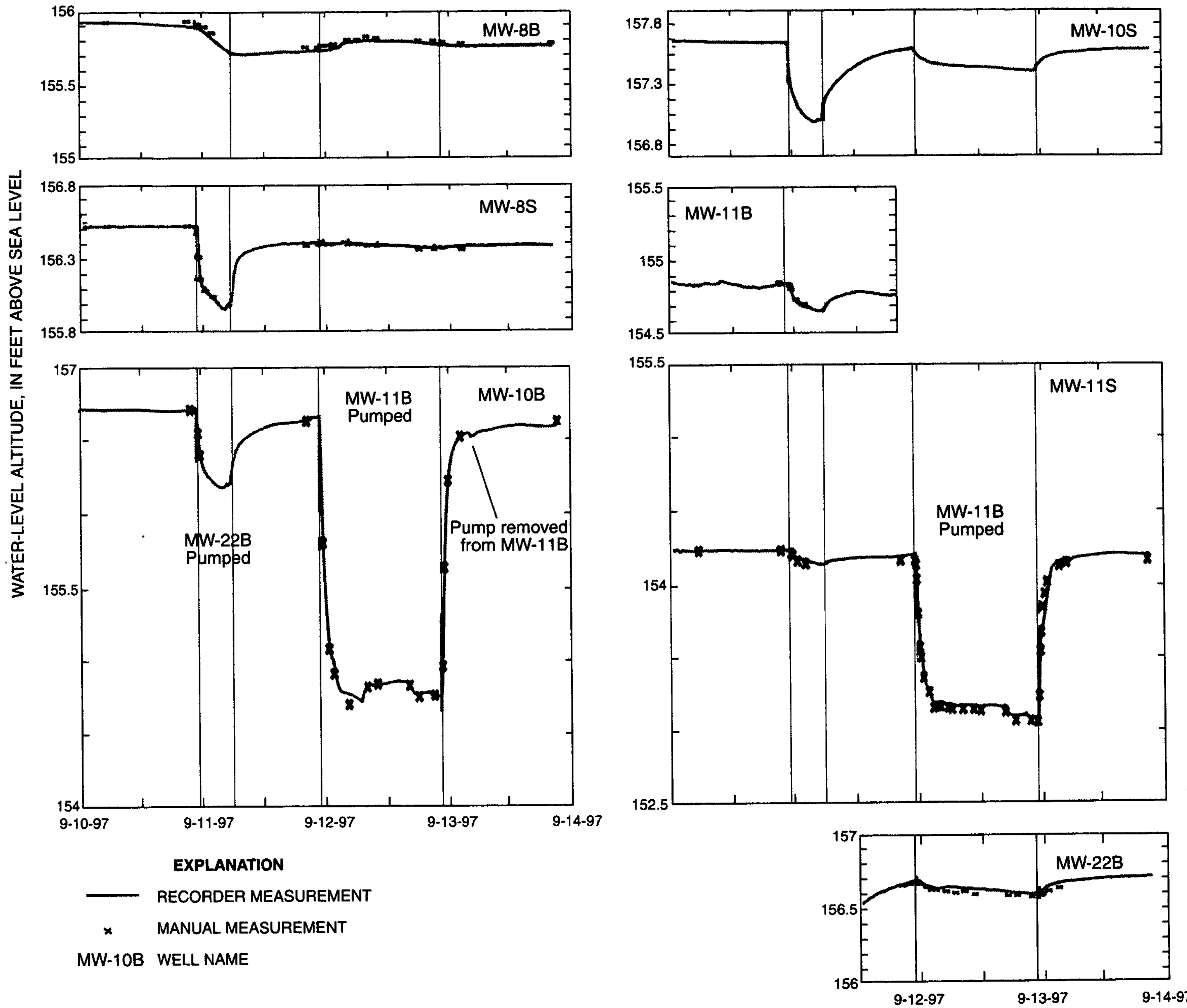
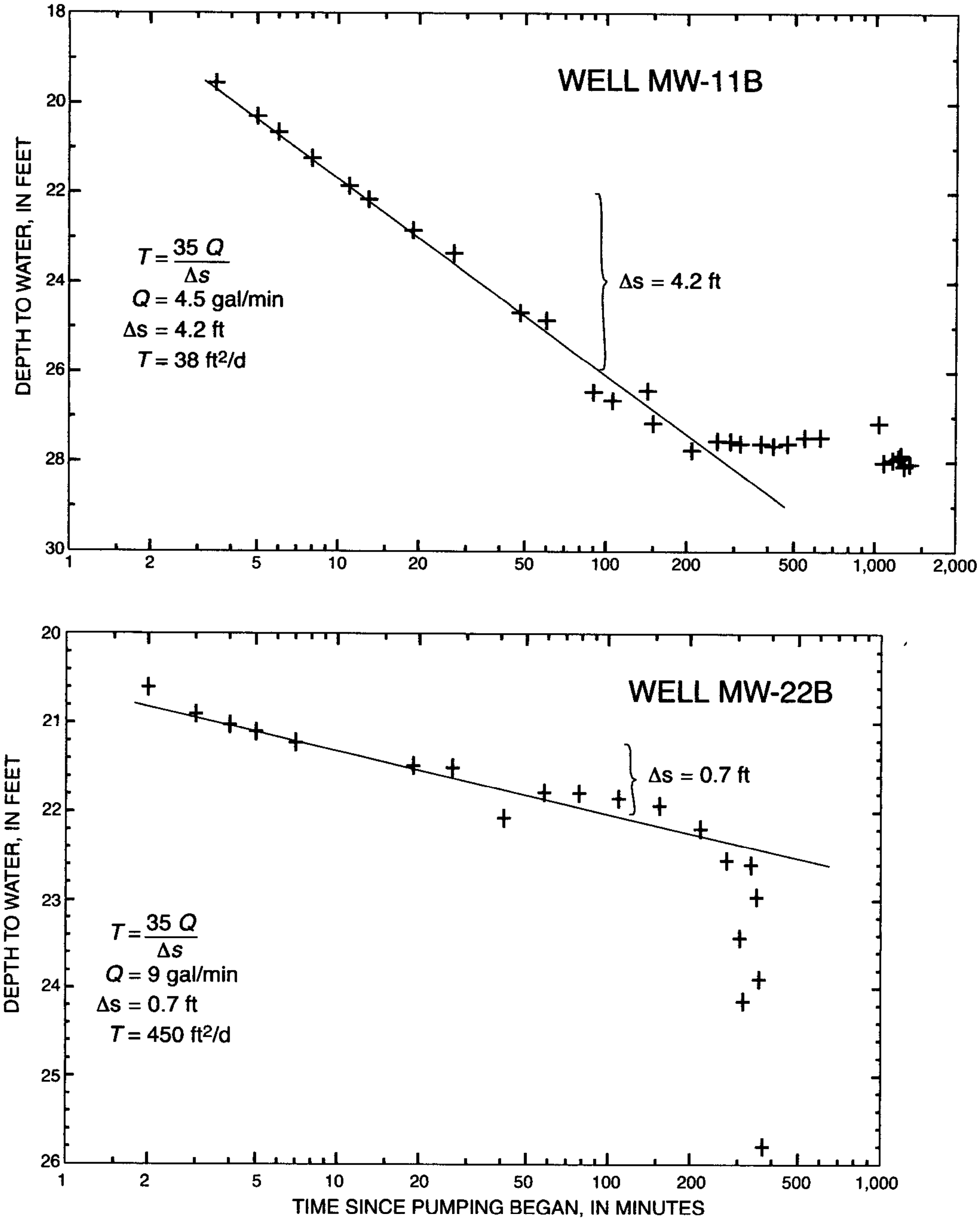


Figure 6. Water-level hydrographs for observation wells that responded to pumping during aquifer testing, Meddybemps, Maine.



EXPLANATION

+ MANUAL WATER-LEVEL MEASUREMENT

——— LINE USED FOR COMPUTATION OF TRANSMISSIVITY

Q PUMPING RATE, IN GALLONS PER MINUTE (gal/min)

Δs CHANGE IN DRAWDOWN, IN FEET (ft) OVER ONE LOG CYCLE

T TRANSMISSIVITY, IN FEET SQUARED PER DAY (ft^2/d)

Figure 7. Depth to water in wells MW-11B and MW-22B during pumping and calculations of transmissivity, Meddybemps, Maine.

ESTIMATION OF HYDRAULIC PROPERTIES USING NUMERICAL METHODS

Numerical methods of analysis were used to simulate conditions in the surficial and bedrock aquifers underlying the study area. Numerical methods have advantages over analytical methods because boundary and initial conditions, along with aquifer properties, can be varied across the area of interest, increasing the degree to which the models represent the conditions found in the aquifer systems. The numerical method used in these analyses is the MODFLOW model, a three-dimensional finite-difference simulator originally documented by McDonald and Harbaugh (1988) and updated by Harbaugh and McDonald (1996).

The use of MODFLOW assumes that ground-water flow in the fractured bedrock underlying the study area can be analyzed (modeled) using a continuum approach. (See Freeze and Cherry, 1979, p. 73, for a discussion of this approach.) This approach involves a representation of the fracture system such that values of hydraulic head and hydraulic conductivity are continuous within the aquifer system and that ground-water flow follows Darcy's Law (that is, flow is proportional to the hydraulic gradient). This approach has been found to be valid and useful in those cases in which a number of open, hydraulically connected fractures are available for ground-water flow at the scale of the model simulation (Domenico and Schwartz, 1990, p. 83–88).

Although it is difficult to assess the degree to which the continuum assumption is valid in the study area, it is clear from borehole-geophysical logs (Hansen and others, 1999) that many water-yielding fractures are present with a spacing between 20 and 50 ft, depending on the orientation of the fracture sets. With this spacing of fractures, it is likely that ground water flows in interconnected sets of fractures across the study area and thus can be simulated as a continuous ground-water-flow system. It should be noted, however, that the hydraulic properties estimated for the bedrock by the continuum approach are average values for blocks of the bedrock, and not of individual fractures or of the non-fractured block. In addition, because of the discrete nature of fractures and their distribution, the hydraulic characteristics of the bedrock system will vary widely, and the local movement of water could deviate from the average movement simulated by the continuum model.

The numerical analysis involved a two-step process, the first of which was to create a simulation of flow in the aquifers at the Meddybemps site under long-term conditions (that is, a steady-state condition). The result from the steady-state simulation then provided a stable initial condition for the second step, which involved transient simulation of the MW-11B aquifer test. These steps were repeated as calibration of the aquifer properties proceeded, with the goal of improving the degree of agreement of simulated heads with measured aquifer heads after each set of changes in aquifer properties made during model calibration. Aquifer properties were changed during model calibration to produce simulated head conditions similar to the heads measured in the aquifer both before the aquifer test (steady-state simulations) and during the aquifer test at well MW-11B (transient simulations).

Model Design and Hydraulic Properties

The design of the Meddybemps numerical model closely follows the description of the ground-water system (conceptual model) presented by Lyford and others (1998) and in the description of the study area in this report. The principal area of interest is the south end of the study area, which includes the southern contaminant plume (fig. 2). The extent of the model (figs. 1 and 8) is much larger than the area of interest so as to minimize the effect of relatively unknown boundary conditions on simulations in the area of interest. To simulate ground-water flow, it was necessary to assign initial conditions, boundary conditions, aquifer properties, and internal sources and sinks of water across the modeled area. A rectangular grid of aquifer blocks was laid out horizontally across the modeled area, and the ground-water flow system was subdivided vertically into layers. The horizontal extent of the grid across the site is shown in figure 8. The modeled area is 40 blocks long along the column direction (northwest to southeast) and 35 blocks wide along the row direction (northeast to southwest). As each block is 50 ft square, the extent of the modeled area is 2,000 ft (column direction) by 1,750 ft (row direction). By convention, the origin in all MODFLOW models is the upper left-hand corner block (column 1, row 1), which in the Meddybemps model is the northernmost corner (fig. 8).

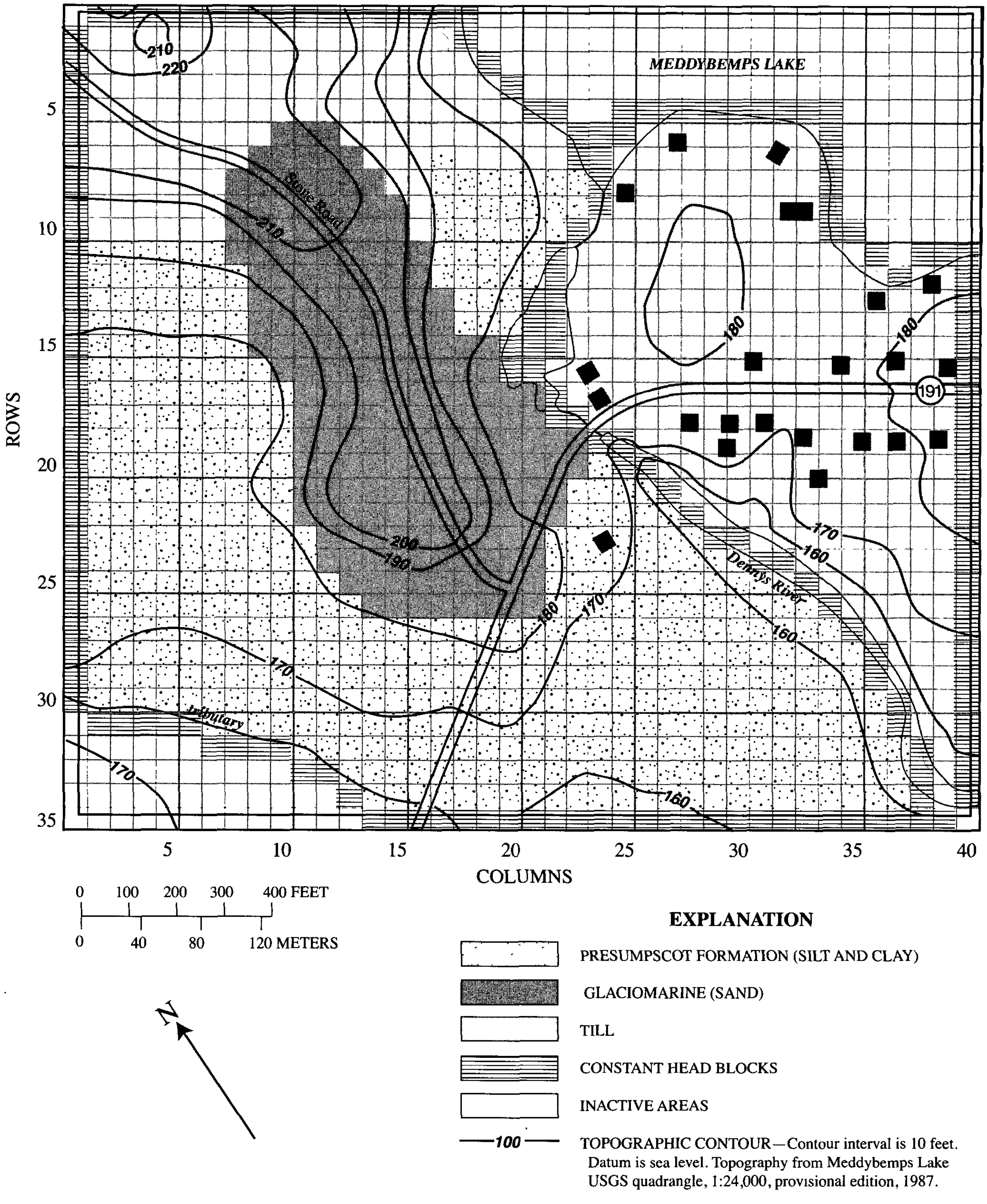


Figure 8. Finite-difference grid, surficial geology of modeled area, and constant-head boundaries, Meddybemps, Maine.

The vertical distribution along section B-B′ (line of section shown on fig. 2) of the four layers used in the Meddybemps model is shown in figure 9. The upper two layers represent the surficial (glacial) materials, with the top layer (model layer 1) representing the uppermost saturated materials. The bottom of layer 1 was established as 5 ft above the bedrock surface. Because most of the surficial aquifer is unconfined, the thickness of layer 1 varied across the modeled area and this layer was inactive in those areas where the saturated thickness of the surficial materials was less than 5 ft. Surficial materials vary across the modeled area as shown in figure 8 and include coarse-grained (sand) glaciomarine deposits on the bedrock ridge to the west of the Dennys River, till, and fine-grained glaciomarine deposits of the Presumpscot Formation west of the river. The hydraulic conductivity of layer 1 varied with the type of surficial material and ranged from 30 ft/d for coarse-grained glaciomarine material, 0.1–0.5 ft/d for till, to 0.001–0.01 ft/d for fine-grained (Presumpscot Formation) material.

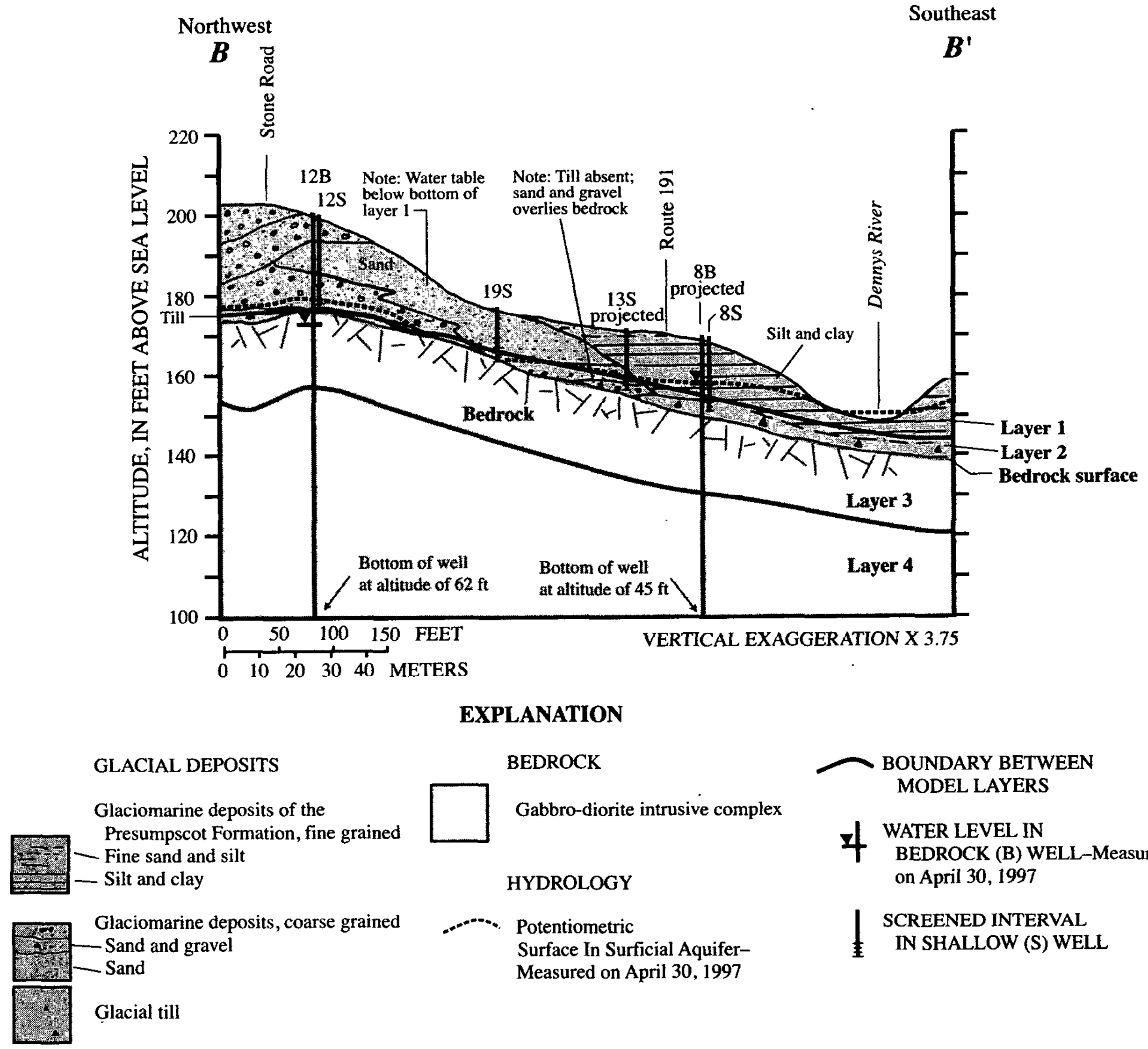


Figure 9. Geologic section B-B′ showing layers used in numerical model, Meddybemps, Maine.

Model layer 2 represents the till that mantles most of the bedrock across the modeled area, although there are significant absences in areas across and adjacent to the bedrock ridge (fig. 9). Layer 2 in those areas represents the lower 5 ft of coarse-grained deposits and was the uppermost active layer in several blocks where the saturated thickness of surficial material was less than 5 ft. The bottom of layer 2 is the bedrock surface across the modeled area. Hydraulic conductivity in layer 2 ranged from 10–30 ft/d for sandy material to 1.0 ft/d for till.

The lower two model layers represent the fractured bedrock aquifer (fig. 9). Layer 3 represents the upper part of the fractured bedrock aquifer, and the bottom of this layer was established at 20 ft below the bedrock surface. Hydraulic conductivity of layer 3 was set almost uniformly to 0.003 ft/d but was increased to 0.5–1.5 ft/d in the area between wells MW-10B and MW-11B because a highly transmissive, high-angle fracture (or fracture set) hydraulically connects these two wells. Layer 4 represents the deeper bedrock aquifer. Because the depth of active ground-water flow in this part of the system in the study area is not well known, the hydraulic properties of this layer were represented using variable transmissivity values. Transmissivity values for layer 4 generally were 0.24 ft^2/d but increased to 40–120 ft^2/d for that area between wells MW-10B and MW-11B with the highly transmissive (high-angle) fracture or fracture set. The combined transmissivity of the two bedrock layers ranged from 0.03 to 150 ft^2/d across the model area, and the majority of the combined bedrock transmissivities were set at 0.3 ft^2/d. A value of 0.3 ft^2/d is consistent with values estimated using specific-capacity data for low-yielding wells (table 2).

Vertical leakage between model layers in MODFLOW is controlled by the VCONT parameter, which is defined as the vertical hydraulic conductivity divided by the distance between vertically adjacent model blocks (thickness); thus VCONT has units of inverse time (day^{-1}). Generally, VCONT values were calculated using equation 51 in the report by McDonald and Harbaugh (1988), which uses the hydraulic conductivity and thickness of the two vertically adjacent blocks. Isotropic conditions were assumed for these calculations (that is, vertical hydraulic conductivity values were held equal to horizontal values). VCONT values between layers 1 and 2 ranged from 4×10^{-4} $days^{-1}$ for fine-grained material to 0.39 $days^{-1}$ for coarse-grained material. VCONT values between layers 2 and 3 were almost uniformly 3×10^{-4} $days^{-1}$. VCONT values between layers 3 and 4 were generally 6×10^{-5} $days^{-1}$. During calibration, VCONT values were increased to 0.005 $days^{-1}$ between layers 2 and 3 and 1.0 $days^{-1}$ between layers 3 and 4 in the region of the high-angle fracture between wells MW-10B and MW-11B, creating a highly transmissive vertical-flow condition in this area of the ground-water flow system. It was also found during model calibration that it was necessary to significantly decrease VCONT values between layers 3 and 4 in the area around the Smith well to match the observed drawdowns in head around this well; VCONT values in this area were 1×10^{-9} $days^{-1}$.

Boundary Conditions

Model boundary conditions include constant-head blocks around the exterior of the modeled area, along the shore of Meddybemps Lake, and along the Dennys River and its unnamed tributary to the south (fig. 8). Head conditions around the model exterior were estimated from surface topography, assuming a 5- to 10-foot depth to ground water. Head conditions for Meddybemps Lake and the Dennys River were based on values reported by Lyford and others (1998) for April 30, 1997. Head conditions along the unnamed tributary to the Dennys River were estimated from surface topography. The constant-head boundary conditions on the exterior of the model, in addition to those on the lake shore and the unnamed tributary on the southern edge of the model, were used in model layers 1 to 4. Constant-head conditions along the Dennys River, transversing the central part of the model, were used only in layers 1 and 2 of the model because the surficial material is absent from most of the streambed.

Recharge and Wells

The source of most ground water underlying the Meddybemps site is recharge from precipitation. Because the rate of infiltration of precipitation is generally controlled by soil permeability, the recharge rates across the study area were varied by the type of material at the surface. An estimated recharge rate of 10 in/yr was used in areas with till at the surface, a rate of 20 in/yr was used in areas with sandy material, and 0.5 in/yr was used in areas underlain by silt and clay

(Presumpscot Formation). The estimates for till and sandy material are consistent with rates reported for similar geologic settings in Connecticut (Mazzaferro and others, 1979). The recharge rate of 20 in/yr that was used in the model is somewhat lower than potential rates of 24 to 26 in/yr (Lyford and Cohen, 1988) to allow for some surface runoff from fairly steep slopes. The value for silt and clay was assumed to be very low, but supporting information is not available.

Two domestic wells were included in the simulation, the Smith well (row 17, column 25) and the Van Wart well (row 18, column 31) (figs. 2 and 8). Both domestic wells were estimated to pump 100 gal/d from the deeper bedrock layer (layer 4) on the basis of steady water-level recovery rates between pumping periods in the Smith well (fig. 5) and a single occupant in the Van Wart residence.

Steady-State Calibration and Simulation

Model simulations of the long-term, steady-state head conditions in the ground-water-flow system underlying the Meddybemps site were conducted to match the geologic and hydrologic observations (calibration), and to provide a stable initial condition for the transient simulations of the MW-11B aquifer test. However, it was found that MODFLOW was highly unstable in the steady-state mode. This instability was the result of the nonlinear nature of the unconfined flow conditions in the model, in which the saturated thickness of the aquifer depends on the head in the model block. This inherently nonlinear situation was made even more unstable in the region of the bedrock ridge near Stone Road (fig. 2) by the large contrast in hydraulic conductivities between the overlying sand and the fractured bedrock below. It was not uncommon during the iterative-solution procedure for the steady-state simulations for wide swings to occur in calculated heads in the region of the bedrock ridge. These swings created high heads in one iteration immediately followed by low heads in a second iteration—an oscillation of calculated head that never allowed the model to converge to a final solution. This model instability occurred in an area of the ground-water system where field measurements indicated that wells that penetrate the surficial materials become dry during some periods of the year, indicating a thin, saturated thickness of the surficial aquifer as an average condition (fig. 9).

The model oscillation problem could not be resolved by adjusting model-solver parameters but was overcome by using a transient simulation that converged to a steady-state condition over a 1,000-day period. The changes in head during the iterative-solution procedure were slow because of the dampening effect of the storage coefficients used in the transient simulation to keep the model solution stable at each time step (0.5 days). The result, after 2,000 time steps, is a condition in which the change in head, with respect to time, has slowed to the point at which the amount of water from storage represents only 1 percent of the total flux through the system, and inputs minus outputs equal the change in storage. The resulting steady-state head solution included several blocks in layer 1 along the bedrock ridge that became inactive (dry) during the simulation, reflecting the small saturated thickness of the surficial material in this part of the ground-water system (fig. 9).

The results of the steady-state simulation are shown in figures 10 and 11. The distribution of simulated head in the uppermost active layer across the modeled area is shown in figure 10; in some areas the uppermost active layer is layer 2, because the calculated head is below the bottom of layer 1. The contoured distribution of head observed in wells also is shown in figure 10; these head contours are modified from Lyford and others (1998, fig. 9). A comparison of the two sets of contours, although limited in areal extent, shows good agreement in absolute magnitude of head and in the direction of ground-water flow. Generally, the primary control on head in the surficial material in the area of interest (that is, the central part of the model) is the movement of ground water from recharge areas in topographically high areas to discharge areas along the Dennys River. To achieve a suitable calibration, the hydraulic conductivity for layer 2 near wells MW-8S, MW-10S, and MW-11S (fig. 2) was increased to a value representative of coarse-grained materials. Low hydraulic conductivity values representative of till were initially assumed because of the low yields observed from wells MW-8S and MW-11S. These wells may not fully penetrate coarse materials, or the coarse materials may not be present near these wells.

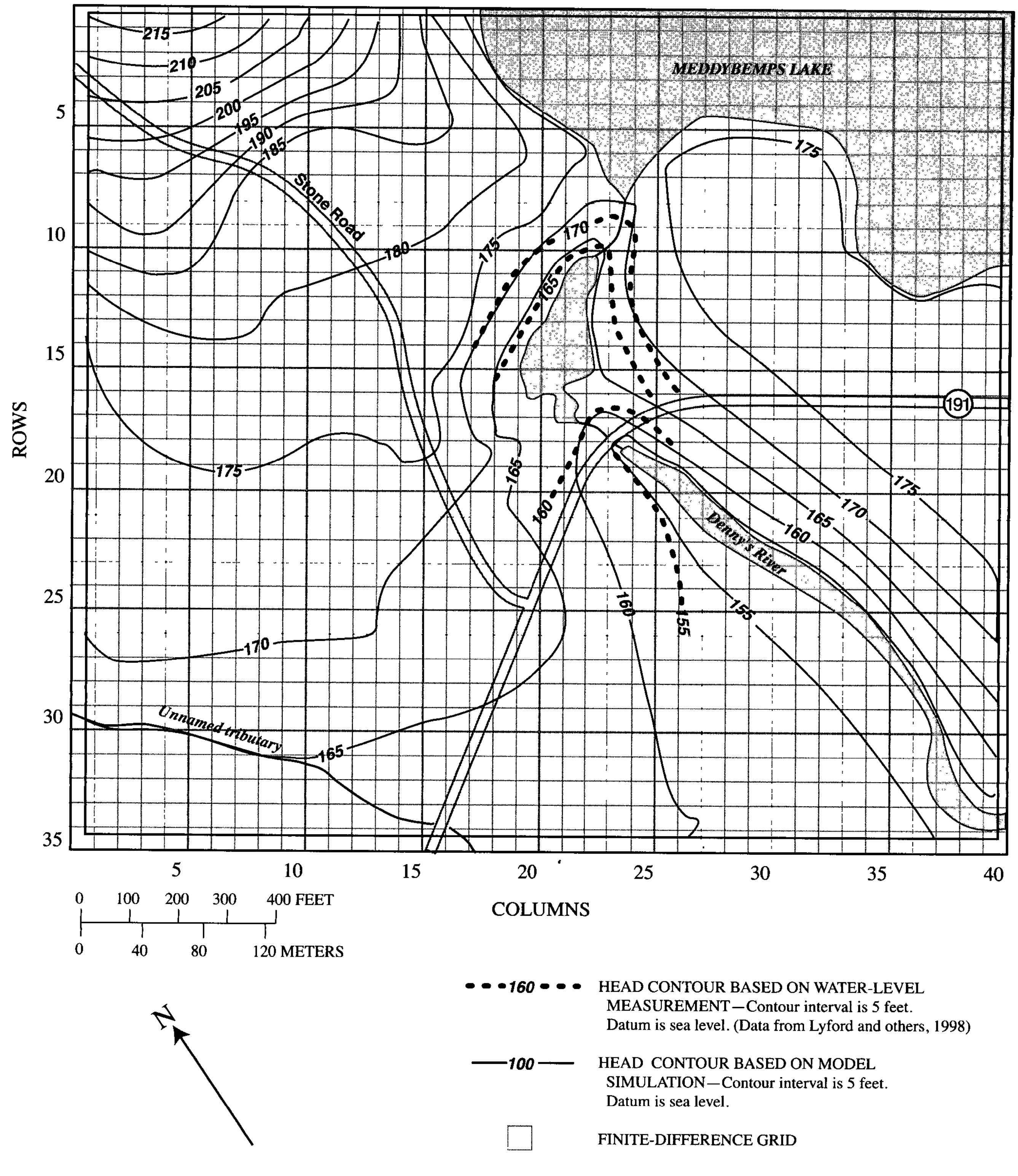


Figure 10. Observed head contours for the surficial aquifer and simulated head contours for the uppermost active layer, Meddybemps, Maine.

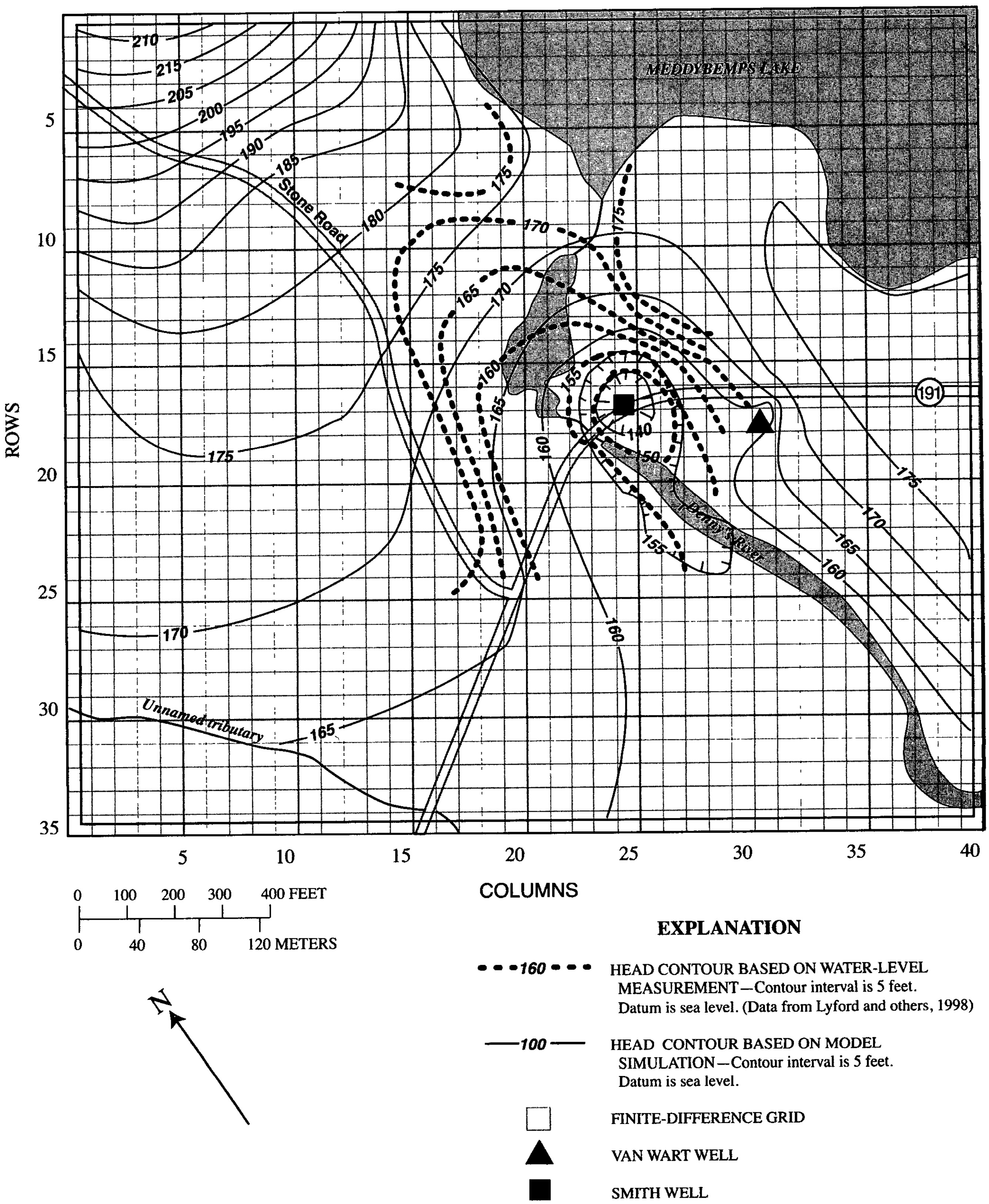


Figure 11. Observed head contours for the bedrock aquifer and simulated head contours for model layer 4, Meddybemps, Maine.

In contrast to conditions found in the surficial aquifer, the primary factor affecting the head in the deeper bedrock (layer 4) in the area of interest appears to be pumping from the Smith well (fig. 11). A well-developed cone of depression is simulated around the Smith well, with a smaller cone developed around the Van Wart well. These simulated head patterns reasonably match the observed head patterns (modified from fig. 11 in Lyford and others, 1998). It should be noted that the vertical leakage parameter, VCONT, was decreased significantly between layers 3 and 4 in the area around the Smith well to simulate the magnitude and extent of the observed head pattern as shown by Lyford and others (1998, fig. 11).

The average difference between the measured and simulated heads (table 4) was about 1 ft, and the standard deviation of those differences was about 4 ft. The largest difference in head was for well MW-1B (fig. 2), a bedrock well on the edge of the area of interest. Most of the differences were small; 85 percent of the differences in head were between -5 and 5 ft. A comparison of well pairs (S-shallow, B-bedrock) indicates that the proper direction of head change with depth was simulated for five of the eight pairs (fig. 2 and table 4). In fact, the largest head difference with depth, measured at wells MW-15S and MW-15B, was closely matched by the simulated head difference with depth at that location. Two well pairs (MW-4S and 4B and MW-7S and 7B) show that measured heads decrease with depth near the Dennys River, whereas the simulated heads increase with depth. This situation may indicate that the effect of drawdown from the Smith well pumpage on heads on the opposite side of the river may be stronger than is simulated by the model.

Table 4. Observed and simulated steady-state heads for wells, Meddybemps, Maine

[Measured head values are water levels, in feet above sea level, on April 30, 1997 (Lyford and others, 1998, table 3)]

Well name	Measured head (h_m)	Simulated head (h_s)	Difference (h_s-h_m)
MW-1B	169.57	179.77	10.20
MW-3B	173.63	170.94	-2.69
MW-4S	165.93	166.01	.08
MW-4B	162.76	169.27	6.51
MW-5S	174.41	175.78	1.37
MW-6S	175.40	179.29	3.89
MW-7S	163.61	163.79	.18
MW-7B	158.79	166.27	7.48
MW-8S	157.93	157.37	-.56
MW-8B	157.01	156.65	-.36
MW-9S	158.43	161.73	3.30
MW-10S	158.90	161.97	3.07
MW-10B	157.90	161.04	3.14
MW-11S	154.99	155.18	.19
MW-11B	155.49	154.95	-.54
MW-12S	177.60	174.90	-2.70
MW-12B	174.22	169.49	-4.73
MW-13S	159.71	159.52	-.19
MW-14B	178.13	183.02	4.89
MW-15S	166.07	168.50	2.43
MW-15B	155.62	158.70	3.08
MW-16S	178.16	172.61	-5.55
MW-16B	176.54	172.53	-4.01
MW-17S	160.57	160.76	.19
MW-18S	160.02	161.79	1.77
MW-20S	175.44	173.36	-2.08
MW-22B	158.36	159.11	.75
Average			1.08
Standard deviation			3.67

Transient Calibration and Simulation of Well MW-11B Aquifer Test

Calibration of model parameters during the transient phase of simulations primarily involved the estimation of aquifer storage properties and changes to aquifer hydraulic properties in the region around wells MW-11B and MW-10B. The type of layer in MODFLOW determines the storage coefficient applied during a simulation. Layer 1 (the uppermost surficial material) was set as a type 1, and an unconfined storage coefficient (specific yield) was used. Layer 2 (surficial material, primarily till) was set as a type 3, or as a convertible layer. If a block in layer 1 is active, the storage coefficient used for the underlying layer 2 is a confined storage coefficient. When the head in a block in layer 1 drops below the bottom of the block, the storage coefficient in the underlying layer 2 block converts to an unconfined storage coefficient. Layers 3 and 4 (bedrock) were set as confined units, and confined storage coefficients were used. Specific yield (unconfined storage) values used in layers 1 and 2 were set at 20 percent. The confined storage coefficient was set at $1x10^{-5}$ for layer 2 and $1x10^{-4}$ for layers 3 and 4. The confined storage coefficient for layer 2 was set

lower than that of layers 3 and 4 because layer 2 is thinner (5 ft) than either layers 3 (20 ft) or 4 (unknown, but several tens of feet at a minimum).

The results of the simulation of the aquifer test are presented in figures 12–14, and table 5. The observed and simulated drawdown over time at observation well MW-10B during the 24-hour test conducted at well MW-11B is shown in figure 12. Simulation time steps ranged logarithmically from 1 to 200 minutes. The shape of the simulated drawdown generally matches the observed drawdown curve. Major features of the observed response in well MW-10B are the rapid increase in drawdown and the equally rapid stabilization of drawdown. To simulate the rapid increase in drawdown over time in the observation well, which is 205 ft from the pumped well, it was necessary to greatly increase the transmissivity of the bedrock aquifer (layers 3 and 4) between these two wells. This change in aquifer properties in the model is supported by geophysical evidence (Hansen and others, 1999), which indicates that a high-angle fracture (or fracture set) hydraulically connects these two wells. In addition to the increase in transmissivity between the two wells, it was also necessary to decrease the transmissivity of the bedrock immediately around the high-transmissivity zone (to

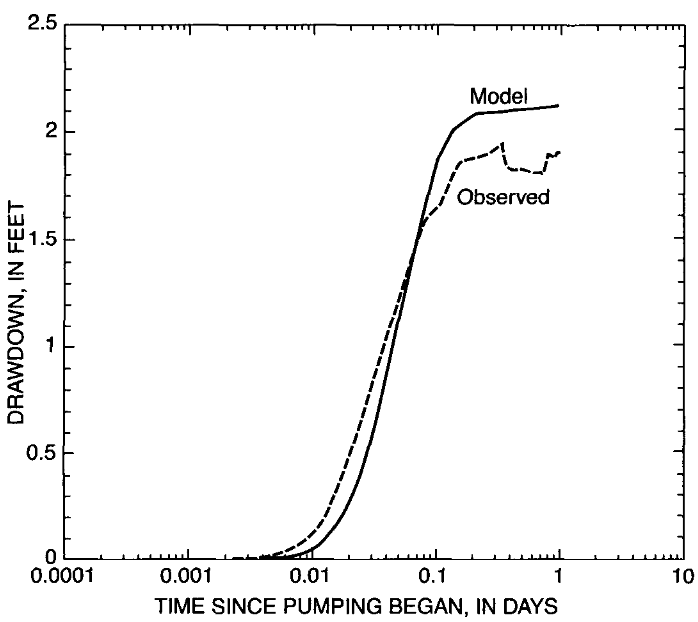


Figure 12. Observed and simulated drawdown at well MW-10B while pumping well MW-11B, Meddybemps, Maine.

0.03 ft^2/d, corresponding to a hydraulic conductivity of $3x10^{-4}$ ft/d) to reduce the drawdown response observed in bedrock wells outside this zone. (See, for example, well MW-22B in table 5.) Although the model was calibrated to simulate the response in MW-22B as if it were open solely to the bedrock system, the low drawdown in this well may be due to vertical flow from the surficial aquifer through a local vertical fracture or a broken casing seal. Therefore, the calibration result of a low hydraulic conductivity zone around the highly transmissive fracture may be an artifact of the questionable observed drawdown at well MW-22B.

The stabilization of the drawdown response in MW-10B (fig. 12) required sources of water to sustain the pumping at MW-11B; the model simulations indicate that the likely sources of water were leakage from the overlying surficial materials and flow from the Dennys River into the bedrock aquifer. A comparison of drawdowns in layer 4 (fig. 13) and layer 2 (fig. 14) shows the effects of these two sources on the drawdown patterns around the pumping well. The steep, elongated cone of depression after 1 day of pumping shown in figure 13 is shortened in the direction of the overlying constant head block in layer 2, indicating the simulated downward leakage of water from the Dennys River into the bedrock fracture that was assumed to extend to the river. The surficial materials are thin along this segment of the Dennys River, providing a potential hydraulic connection between the river and the bedrock fracture, if it underlies the river. In addition to leakage from the river, the model simulations also indicated that water leaking downward from the overlying surficial material into the fracture was needed to stabilize drawdown induced by pumping of MW-11B. Because the dip angle of the fracture is high (>45 degrees), the outcrop of the fracture at the bedrock surface is expected to be present near the pumping and observation wells, and the vertical leakage parameter VCONT was increased to 0.005 $days^{-1}$ between layers 2 and 3 and 1.0 $days^{-1}$ between layers 3 and 4 to simulate this ability to move water vertically.

The drawdown pattern in the surficial aquifer (layer 2) after 1 day of pumping is shown in figure 14. Although the drawdowns in layer 2 were significantly less than those in the fracture (fig. 13), they are areally extensive and indicate that water was moving through the overlying materials toward and into the fracture, providing another source of water for drawdown stabilization. As noted earlier, the hydraulic

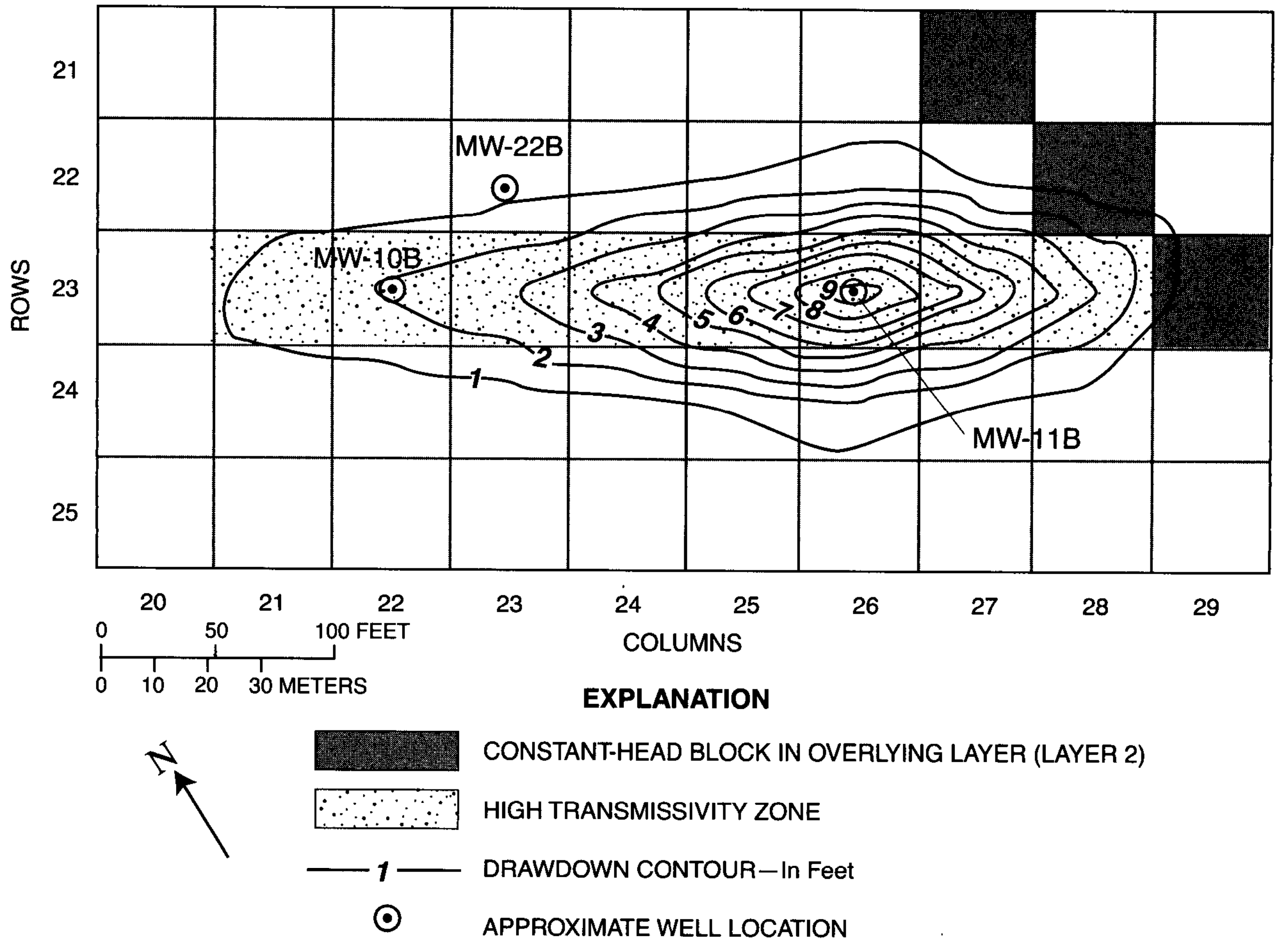


Figure 13. Simulated drawdown in model layer 4 after pumping MW-11B for 1 day, Meddybemps, Maine.

conductivity of the surficial aquifer in this area (layer 2) was increased from initial estimates to simulate the drawdowns observed in these wells (table 5) and to improve the steady-state head results with respect to pre-test conditions.

The overall magnitude of simulated drawdown values was reasonable for the five observation wells and the pumped well at the end of the aquifer test (table 5). Differences between observed and simulated drawdowns at the observation wells can be explained, to some degree, by differences in the distances between observed and simulated wells (table 5). Additionally, it should also be recognized that the drawdown value in the pumped well (MW-11B) is expected to be larger than that of the simulated drawdown in the block containing the pumped well, because the numerical model assumes a distributed pumping source (that is, the water is pumped from all of the 50-ft square block). Head losses at the open well bore (also called entry losses) are considered to be negligible at the pumping rate of 4.5 gal/min. It is particularly interesting to note the greatly reduced drawdown in well MW-22B relative to MW-10B, although it is significantly closer to the pumped well than MW-10B (fig. 2). However, water levels in MW-22B may not be suitable for comparison to modeling results for reasons stated earlier. In addition, although not shown in table 5, it should be noted that no drawdown was observable in bedrock well MW-8B during the aquifer test, even though 0.04 ft of drawdown was recorded in the shallow well at that location. These observations support the view that the major control on the flow of water in this region of the bedrock system is the highly transmissive fracture set between wells MW-11B and MW-10B.

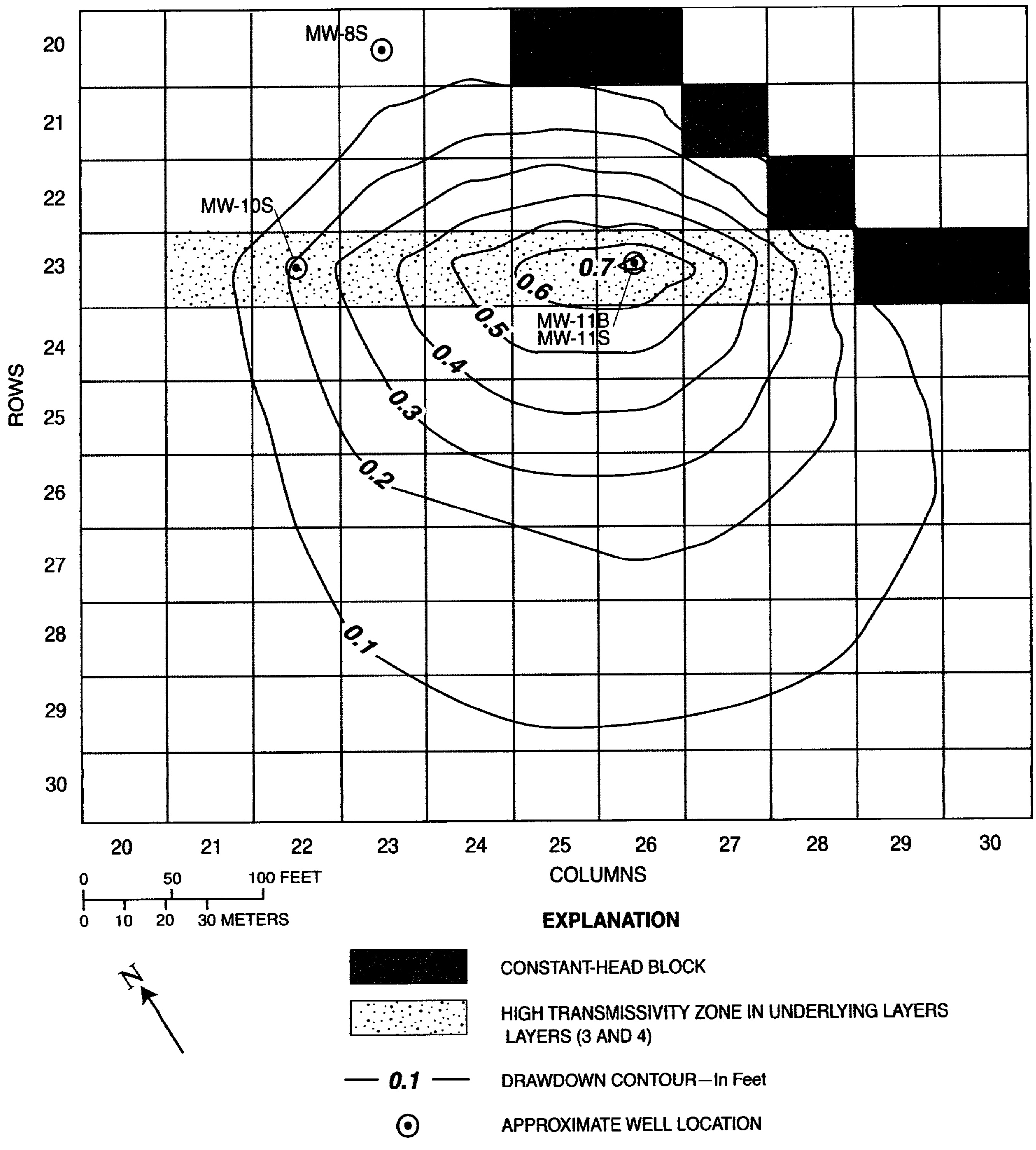


Figure 14. Simulated drawdown in model layer 2 after pumping MW-11B for 1 day, Meddybemps, Maine.

Table 5. Observed and simulated drawdown in the pumped well and observation wells after pumping well MW-11B for 24 hours, Meddybemps, Maine

Well name	Distance to pumped well (feet)	Observed drawdown (feet)	Simulation distance (feet)	Simulated drawdown (feet)	Difference (simulated-observed) (feet)
MW-8S	202	0.04	212	0.03	-0.01
MW-10S	210	.16	200	.22	.06
MW-10B	205	1.90	200	2.12	.22
MW-11S	15	1.12	12	.71	-.41
MW-11B	0	11.78	12	9.91	-1.87
MW-22B	169	.10	158	.45	.35

SUMMARY AND CONCLUSIONS

The hydraulic conductivity of surficial materials near the Eastern Surplus Superfund Site in Meddybemps, Maine, determined from specific-capacity tests, ranges from 17 to 78 ft/d for wells completed in coarse-grained glaciomarine sediments and from about 0.1 to 1.0 ft/d for wells completed in till. The transmissivity of fractured bedrock determined from specific-capacity tests and an aquifer test in bedrock ranges from about 0.09 to 130 ft^2/d. Relatively high values of transmissivity at the southern end of the study area appear to be associated with a high-angle fracture or fracture zone that hydraulically connects two wells completed in bedrock. Transmissivities at six low-yielding (less than 0.5 gal/min) wells, which appear to lie within a poorly transmissive block of the bedrock, were consistently in a range of about 0.09 to 0.5 ft^2/d.

The estimates of hydraulic conductivity and transmissivity in the southern half of the study area were supported by steady-state calibration of a numerical model and simulation of a 24-hour aquifer test at a well completed in bedrock. Model calibration indicated an extension of a relatively transmissive zone in the surficial aquifer beyond the mapped extent of coarse-grained sediments eastward to the Dennys River. It was necessary to use high values of vertical hydraulic conductivity along the fracture or fracture zone that connects the pumped well and an observation well to match drawdowns during model calibration to the aquifer tests. In addition, modeling results indicated that a very low vertical hydraulic conductivity was needed to simulate a persistent cone of depression near a residential well that may lie within the poorly transmissive block of bedrock.

Considerable uncertainty is attached to estimates of transmissivity and hydraulic conductivity determined by analytical methods because of the many simplifying assumptions. Nevertheless, calibration of a numerical model yielded values of the hydraulic properties of the surficial- and bedrock-aquifer system that were similar to those determined by analytical methods for wells at the Meddybemps site. Numerical modeling also yielded estimates of vertical hydraulic conductivity that would be difficult to determine analytically from available data. Generally, the numerical modeling results support the conceptual model of ground-water flow. The following observations were based on numerical modeling:

- Hydraulic properties within the surficial materials and the fractured bedrock are quite variable. Hydraulic conductivities range between 0.001 to 30 ft/d for surficial materials, and transmissivities range from 0.03 to 150 ft^2/d for fractured bedrock. The high contrast in values in hydraulic properties indicate the presence of preferential pathways for water flow in the Meddybemps site that could affect the direction and rate of contaminant transport at the site.
- Although low yields in wells MW-8S and MW-11S indicate that the transmissivity of surficial materials decreases eastward and southeastward toward the Dennys River, model calibration indicated that the surficial materials in this area are relatively transmissive.

- The rapid stabilization of the drawdown response to pumping at well MW-11B indicates that likely sources of water are leakage from the overlying surficial materials and flow from the Dennys River into the bedrock aquifer. Simulation of downward movement of water from surficial materials and lateral movement from the Dennys River to the principal fracture zone yielding water in well MW-11B produced asymmetric and elliptical drawdown patterns in bedrock and surficial aquifers.
- Model simulations indicate that the vertical leakage to the deeper bedrock around the Smith residential well must be very low to maintain the depressed water levels that are observed in nearby bedrock wells and in the Smith well itself. The low vertical leakage and the low transmissivity of the deeper bedrock in this part of the study area is supported by the specific-capacity data and head observations for wells open to this part of the bedrock. The inferred low vertical leakage, in combination with the low transmissivity of the bedrock in this part of the study area, could constrain contaminant transport. VOCs had not been detected in the Smith well prior to the spring of 1998. The high degree of variability in aquifer hydraulic properties, however, must be taken into consideration to achieve an accurate assessment of the vulnerability of an individual well to contamination.

The estimates of hydraulic properties presented in this report were determined on the basis of available specific-capacity data, aquifer tests, and numerical modeling done within the time and budgetary constraints of the project. Additional aquifer testing and further refinement of the model through additional calibration might narrow the range of hydraulic property values and refine the conceptual flow model for the aquifer system; however, the degree of refinement needed will depend on the intended use of the information. The information presented in this report could be used to estimate ground-water-flow patterns and flow rates of water through the aquifer system and to assess remediation approaches. Design of a ground-water remediation system and simulation of solute transport might require additional information, particularly the hydraulic properties of fractures.

The numerical model described in this report was designed to analyze aquifer tests, evaluate the conceptual model of ground-water flow, and refine estimates of hydraulic properties for the aquifer system. Other uses, such as a design of a ground-water remediation system or as a basis for solute-transport modeling, may not be appropriate because fracture distribution and properties have been characterized only in some areas and other conceptual models of ground-water flow may be possible. The preponderance of evidence, however, indicates that the conceptual model is the most likely interpretation of information collected thus far at the Meddybemps site.

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